NASA TECHNICAL TRANSLATION

NASA TT F-328

NASA TT F-328

N65-21436

(ACCESSION NUMBER)

10

(PAGES)

(NASA CR OR TMX OR AD NUMBER)

(THRU)

1

(CODE)

13

(CATEGORY)

FACILITY FORM 602

SPECTRAL REFLECTIVITY OF CLOUDS

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Hard copy (HC) ______

Microfiche (MF) ______

From *Trudy Glavnoy Geofizicheskoy Observatorii imeni A. I. Voyeykova,* No. 152, 1964

NATIONAL AERONAUTICS AND SPACE ADMINISTRATION • WASHINGTON, D. C. • APRIL 1965

NASA TT F-328

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Translation of "Spektral'naya otrazhatel'naya sposobnost' oblakov." Trudy Glavnoy Geofizicheskoy Observatorii imeni A. I. Voyeykova, No. 152, pp. 186-191, 1964.

NATIONAL AERONAUTICS AND SPACE ADMINISTRATION

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SPECTRAL REFLECTIVITY OF CLOUDS

Ye. P. Novosel'tsev

ABSTRACT

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The article considers the question of the spectral reflectivity of waterdrop clouds. It is shown that the clouds remain practically "gray" up to $\lambda \approx 0.7\ \mu$. In the absorption bands of water vapor, some decrease in reflectivity is observed For $\lambda > 2.5\,\mu$, due to a large increase in the absorption of liquid water radiation, there is a sharp drop in the value of the cloud's albedo. A method is proposed for an approximate evaluation of the spectral albedo of ice clouds.

The investigation of the spectral properties of these clouds is of great interest in connection with a whole series of meteorological problems as well as problems of aerial photography. However, because of technical difficulties, direct experimental determination of the spectral albedo of clouds has not been realized to date. Therefore, it is particularly important to examine this problem theoretically. We should point out that a theoretical solution will make it possible to determine the albedo of clouds for any type of base surface whose spectral properties control, to a large extent, the spectral albedo of the clouds (this applies particularly to clouds of the upper layer).

If the formulated problem can be solved for waterdrop clouds with sufficient reliability, for ice clouds we can obtain only very rough quantities, merely giving a quantitative picture of the true variation in the spectral reflectivity of clouds.

We will first consider the case of the spectral albedo of waterdrop clouds of sufficient power (i.e., clouds of the lower and medium layers).

According to the results of reference 1, the system of equations describing the transmission of radiation in a scattering and an absorbing medium has the following form

$$\left.\begin{aligned}\frac{dF^{(1)}}{d\tau} &= -m^{(1)}(\tau)\left[\sigma\Gamma(\tau)+K\right]F^{(1)}(\tau)+m^{(2)}(\tau)\,\sigma\Gamma^{(2)}(\tau)\,F^{(2)}(\tau),\\ \frac{dF^{(2)}}{d\tau} &= m^{(1)}(\tau)\,\sigma\Gamma^{(1)}(\tau)\,F^{((1)}(\tau)-m^{(2)}(\tau)\left[\sigma\Gamma^{(2)}(\tau)+K\right]F^{(2)}(\tau),\end{aligned}\right\} \qquad (1)$$

where $F^{(1)}(\tau)$ and $F^{(2)}(\tau)$ are the ascending and descending fluxes, τ is the optical thickness, σ is the volumetric coefficient of scattering, K is the volumetric coefficient of absorption, and $m^{(1)}$ and $m^{(2)}$ are the average secants of the zenith angles determining the direction of propagation of the "centers of gravity" of the ascending and descending fluxes.

In the case of thick clouds we assume that $m^{(1)} = m^{(2)} = m_D$ ($m_D \approx 2$); $\Gamma^{(1)}$ and $\Gamma^{(2)}$ determine, respectively, the portion of the ascending and descending radiation scattered by the elements of volume in the upper or lower hemisphere. For thick clouds $\Gamma^{(1)} = \Gamma^{(2)} = \Gamma_D$ ($\Gamma_D \approx 0.05$) (ref. 2).

The system of equations (1) is solved for the following boundary conditions: at the upper boundary of the clouds $F^{(2)}(\tau^*) = S \sin h_\odot$, where S is the incident flux of solar radiation, $h_\odot$ is the angular altitude of the sun above the horizon; at the surface of the Earth $F^{(1)}(0) = F^{(2)}(0)\, a$, where a is the value of the base surface albedo.

In considering the question of the albedo of thick clouds, we may neglect the effect of layers above and below the clouds.

By solving the system of equations (1) with given boundary conditions we obtain the following expression for the value of the cloud albedo:

$$a(t^*) = \frac{[(1-a)+\beta(1+a)]\,e^{\alpha t^*}(1-\beta) - [(1-a)-\beta(1+a)]\,e^{-\alpha t^*}(1+\beta)}{[(1-a)+\beta(1+a)]\,e^{\alpha t^*}(1+\beta) - [(1-a)-\beta(1+a)]\,e^{-\alpha t^*}(1-\beta)}, \tag{2}$$

where $\alpha = m_D\sqrt{1-g^2}$, $\beta = \sqrt{\frac{1-g}{1+g}}$, $g = \frac{\sigma\Gamma_D}{\sigma\Gamma_D + K}$, $t^* = \int_0^{z^*}(\sigma\Gamma_D + K)\,dz$,

z* being the geometric thickness of the cloud.

For determining the spectral albedo we require reliable data about the coefficients which enter into expression (2). We consider the question of these coefficients in a little more detail.

The absorption coefficient K consists of the sum of two coefficients: the absorption coefficient of radiation by cloud droplets k_{cd} and the absorption coefficient of radiation by water vapor k_{wv}.

The magnitude of the volumetric absorption coefficient k_{cd} can be determined from the following equation

$$K_{cd} = \frac{3}{4}\frac{\rho}{\bar{r}}(1-\bar{R})\left(1-e^{-2\alpha'\bar{r}}\right),$$

where $\bar{r}$ is the average radius of the cloud droplets, β is the average liquid water content of the clouds (for clouds of the lower and medium layers this quantity is equal to 8 μ, on the average), α' is the absorption coefficient of liquid water, and $\bar{R}$ is the coefficient which determines the value of radiation reflected by the droplets.

In the interval of wavelengths 0.4-2.5 μ, $\bar{R}$ varies within the limits 0.06 - 0.07.

For clouds of the middle layer, ρ is assumed to be equal to 0.25×10^{-6} g/cm^3, and for clouds of the lower layer it is equal to 0.35×10^{-6} g/cm^3 (refs. 3 and 4).

The coefficient σ is obtained from the differences $S-k_{cd}$, where S is the volumetric attenuation coefficient. The quantity S is obtained from the following relationship (ref. 5)

$$S = \frac{3}{2}\frac{\rho}{\bar{r}}.$$

The volumetric coefficients k_{cd} and σ can, therefore, be obtained in a rather simple manner. The situation is more complicated when we determine the absorption coefficient of radiation by vapor k_{wv}.

As we know, we can use the concept of absorption coefficient only when Buger's law is satisfied. However, when radiation is absorbed by water vapor, Buger's law is not satisfied even for the case of very narrow spectral intervals. Consequently, when radiation is absorbed by water vapor, the concept of the absorption coefficient has no meaning, and in this case we must deal with the transmission functions.

Ye. S. Kuznetsov analyzed experimental data and proposed the following empirical equation for the transmission function

$$P(w) = \gamma e^{-cw} + (1-\gamma)e^{-dw},$$

which is quite well justified in a wide range of wavelengths.

In this equation w is the water vapor content, c and d can be interpreted as the average mass absorption coefficients at two regions of the spectral interval over which the given absorption band is broken down in some fashion, while γ and $(1 - \gamma)$ can be interpreted as the relative lengths of these regions.

In this case the large number of points characterized by coefficient c are not necessarily associated (ref. 6). The form of the transmission function proposed by Kuznetsov is convenient, because it permits us to use the concept of the absorption coefficient. Indeed, if we break down the absorption band into regions γ and $1 - \gamma$, we may assume for the first region $k_{wv} = cw_0$, while for the second region $k_{wv} = dw_0$, where w_0 is the absolute humidity determined from the well-known Magnus equation.

The values of the quantities c, d and γ for six basic absorption radiation bands of water vapor are shown in reference 6.

In solving this system of equations (1), it was assumed that the coefficient $g = \frac{\sigma\Gamma_D}{\sigma\Gamma_D + k_{cd} + k_{wv}}$ does not depend on the altitude above the lower boundary of the cloud.

At the present time, we do not have sufficiently valuable data to investigate in detail the question of the variation in parameter g with altitude. On the other hand, a qualitative evaluation shows that our proposition that coefficient does not depend on altitude is quite justified.

As an illustration, we computed the spectral albedo of the clouds of the middle layer (the geometric thickness of the clouds of the medium layer is 500 m).

The calculations were carried out as follows: the entire spectral range consisting of shortwave solar radiation was broken down into intervals, each of which was equal either to the width of one of the absorption bands of water vapor or to the width of the interval between them. For the interval including the absorption band, the albedo was determined as an average value in accordance with the following equation

$$a = a_c\gamma + (1 - \gamma)\,a_d,$$

where a_c is the albedo corresponding to region γ, and a_d is the albedo corresponding to region $(1 - \gamma)$.

We are not able to determine the localization of these regions. Therefore, the spectral "resolution" (if we use the instrument terminology) of our method in the water vapor absorption bands is not very high. In the intervals between the water vapor absorption bands the "resolution" may be sufficiently high, because only the liquid water with a continuous spectrum of absorption absorbs in this case.

In the calculations the base surface was assumed to be "gray" with $a_{base} = 0.2$.

The relationship $a(\lambda)$ which we obtained is shown in figure 1.

As we can see from the figure, the albedo of the cloud does not depend on the wavelength up to $\lambda \approx 0.7\ \mu$. The dips in the curve are explained basically by the absorption of radiation by water vapor. When $\lambda > 2.5\ \mu$, the absorption by liquid water increases sharply, so that when $\lambda = 3\ \mu$, the clouds become almost black.

As we have already pointed out in the case of ice clouds of the upper layer, we can make only a very rough evaluation of their spectral reflectivity, because at the present time there are no reliable data concerning the optical characteristics of these clouds.

Let us proceed with a formal solution of our problem. The analytic expression for the value of the albedo of the upper layer clouds may be represented, approximately, in a form analogous to (2). However, the relationships $m^{(1)} = m^{(2)} = m_D$ and $\Gamma^{(1)} = \Gamma^{(2)} = \Gamma_D$, which are valid only in the case of very large optical thickness, will not be applicable here.

In the case of small optical thickness, (the optical thickness of the clouds of the upper layer), the solution of system (1) is transformed into another limiting case:

$$a(t^*) = \frac{[(1-a_c)+\nu(1+a_c)]\,e^{\mu t}\,(1-\nu) - [(1-a_c)-\nu(1+a_c)]\,e^{-\mu t^*}\,(1+\nu)}{[(1-a_c)+\nu(1+a_c)]\,e^{\mu t^*}\,(1+\nu) - [(1-a_c)-\nu(1+a_c)]\,e^{-\mu t^*}\,(1-\nu)}, \qquad (2')$$

where

$$\nu = m_{\odot}\sqrt{1-q^2}, \qquad \mu = \sqrt{\frac{1-q}{1+q}}, \quad q = \frac{\sigma\Gamma_{\odot}}{\sigma\Gamma_{\odot}+K}.$$

Here $m_{\odot}$ is the secant of the zenith angle of the sun, $\Gamma_{\odot}$ is the portion of the direct radiation scattered by the elementary volume in the upper hemisphere, (in accordance with reference 7 $\Gamma_{\odot} \approx 0.25$), a_c is the spectral albedo of the system-base surface-atmosphere below the clouds, $t^* = \int_0^{z^*} (\sigma\Gamma_{\odot}+K)\,dz$, σ is the volumetric scattering coefficient, and $K = k_{cd} + k_{wv}$ (k_{cd} is the volumetric absorption coefficient for ice crystals).

Since the attenuation of the radiation in the clouds of the upper layer takes place almost entirely because of scattering by large particles (the role played by absorption is small), t^* is practically independent of wavelength.

In order to compute the spectral albedo of ice clouds, we must know the following three quantities: a_c, t^* and q. The value of a_c can be determined with sufficient reliability if we know the spectral albedo of the base surface.

The quantity t^* can also be determined with sufficient reliability (ref. 7). In order to compute the spectral albedo of the clouds of the upper layer, we must determine quantity q. Let us assume that we are dealing with clouds whose optical thickness is infinite, and which consist of ice crystals. In this case we quite easily obtain the following expression from (2), i.e., the albedo of such a cloud depends on the magnitude of parameter g.

$$a(\lambda)=\frac{\sqrt{1-g(\lambda)}-\sqrt{1-g(\lambda)}}{\sqrt{1+g(\lambda)}-\sqrt{1+g(\lambda)}},$$

If we were to measure the spectral albedo of an ice cloud of infinite thickness, we would determine the value of the parameter

$$g(\lambda)=\frac{2a(\lambda)}{1+a^2(\lambda)}.$$

Strictly speaking, parameter g which enters into expression (2), obtained for the case of layers with large optical thickness and parameter q in expression (2') may be somewhat different. However, because in the first approximation for the case of ice clouds we assume that $\Gamma_{\odot}=\Gamma_D\approx 0.25$, we have $g(\lambda)\approx q(\lambda)$.

A rough analog of such a cloud with infinite optical thickness may be a sufficiently thick layer of dry, newly fallen, fluffy snow. However, it is necessary to point out that solution (2'), strictly speaking, is not applicable to the case of radiation propagation in snow, because with this concentration of particles we must deal with fields and not with intensities. The transmission equation, from which the system of equations (1) was obtained, was formulated for intensities. However, for qualitative evaluation this method is apparently applicable.

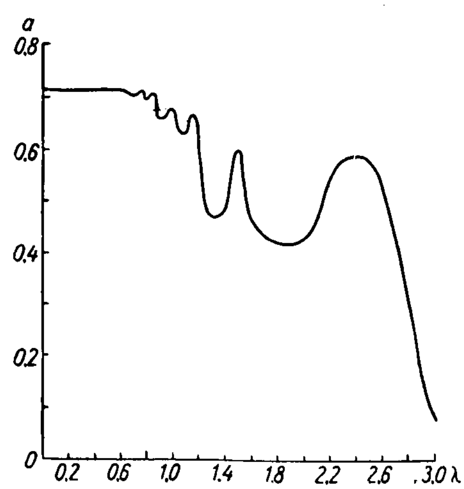

Figure 1. The spectral albedo for clouds of the middle layer.

Very thorough measurements of the spectral albedo of snow were carried out by L. B. Krasil'shchikov. By using the data (kindly presented by Krasil'shchikov) on the spectral albedo of fresh dry snow we obtain the following values for parameter g (Table 1).

However, the optical properties of snow are determined exclusively by the optical properties of crystals, while the optical properties of clouds to some degree (although a small degree) depend on the optical properties of water vapor. Therefore, it is necessary to determine not

$$q = \frac{\sigma\Gamma}{\sigma\Gamma + k_\lambda}, \tag{3}$$

but

$$q_1 = \frac{\sigma\Gamma}{\sigma\Gamma + k_\lambda + k_{WV}}. \tag{4}$$

From (3) and (4) is is easy to obtain $q_1 = \dfrac{q}{1 + q_1 \dfrac{k_{WV}}{\sigma\Gamma_\odot}}$.

TABLE 1.

λ	0,70	0,75	0,80	0,85	0,90	0,95	1,00	1,05	1,10	1,15
g	0,99	0,995	1,00	1,00	1,00	0,998	0,994	0,985	0,991	0,967
λ	1,20	1,25	1,30	1,35	1,40	1,45	1,50	1,55	1,60	1,65
g	0,883	0,873	0,875	0,834	0,640	0,260	0,100	0,100	0,174	0,293
λ	1,70	1,75	1,80	1,85	1,90	1,95	2,00	2,05	2,10	2,15
g	0,362	0,400	0,400	0,236	0,050	0,00	0,00	0,074	0,246	0,269
λ	2,20	2,25	2,30	2,35	2,40	2,45	2,50			
g	0,269	0,246	0,149	0,198	0,024	0,024	0,00			

Note: Commas in this table represent decimal points.

Consequently, it is also necessary to evaluate the ratio $\dfrac{k_{WV}}{\sigma\Gamma_\odot}$ in order to determine the quantity q_1.

The quantity k_{WV} can be determined quite simply from the temperature of the clouds in conjunction with the methodology presented above (for the case of waterdrop clouds).

The value of $\Gamma_\odot \approx 0.25$, $\sigma = 0.5 \cdot 10^{-5}$ (ref. 7).

All quantities necessary for the calculations can be determined, and the calculation of the spectral albedo of ice clouds can be carried out by means of equation (2').

In conclusion we note that when the ice clouds are observed from the top, the spectral composition of the ascending radiation flux will depend not only on the spectral albedo of the clouds (since these clouds are sufficiently thin and transparent), but also on the spectral albedo of the base surface. Therefore, the calculation based on the assumption of a "gray" base surface and a "gray" atmosphere below the clouds does not give a correct picture of the spectral composition of the radiation from the cloud.

REFERENCES

1. Kuznetsov, Ye. S. The General Method of Constructing Approximate Equations for Transmission of Radiant Energy (Obshchiy metod postroyeniya priblizhennykh uravneniy perenosa luchistoy energii). Izv. AN SSSR, ser. geogr. i geofiz., No. 4, 1951.

2. Berlyand, M. Ye. and Novosel'tsev, Ye. P. The Theory of the Variation in the Total Radiation with Cloudiness (K teorii zavisimosti summarnoy radiatsii ot oblachnosti). Nauch. soobshch. In-ta geologii i geografii AN LitSSR, Vol. 13, 1962.

3. Fizika Oblakov. The Physics of Clouds. Edited by A. Kh. Khrgian. Gidrometeoizdat, Leningrad, 1961.

4. Minervin, V. Ye., Mazin, I. P. and Burkovskaya, S. N. Some New Data on the Liquid Water Content of Clouds (Nekotoryye novyye dannyye o vodnosti oblakov). Trudy TsAO, No. 19, 1958.

5. Shifrin, K. S. Scattering of Light in a Turbid Medium (Rasseyaniye sveta v mutnoy srede). Gostekhizdat, Moscow-Leningrad,1951.

6. Kuznetsov, Ye. S. Absorption of Solar Radiation by the Earth's Atmosphere (O pogloshchenii radiatsii solntsa zemnoy atmosferoy). Trudy Geofiz. in-ta., No. 23 (150), 1954.

7. Novosel'tsev,Ye. P. Liquid Water Content of the Upper Layer (O vodnosti oblakov verkhnego yarusa). Meteorologiya i godrologiya, No. 8, 1962.

Translated for the National Aeronautics and Space Administration
by John F. Holman and Co. Inc.

NASA-Langley, 1965 F-328